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Coastal Flow in the Northwest Gulf of Alaska: The Kenai Current

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Recent data from the northwest Gulf of Alaska reveal a coastal current which flows westward along the Kenai Peninsula (mainly within 30 km of shore), enters Shelikof Strait, and exits to the southwest of Kodiak Island. This flow, which we call the Kenai Current, has a large seasonal variation in baroclinic transport and maximum surface speed; transport is typically about 0.3×10^6 m^3/s but exceeds 1.0×10^6 m^3/s in fall, with concurrent speed increases from 15–30 cm/s to over 100 cm/s. The coastal flow is clearly distinct from the offshore Alaskan Stream; its seasonal signal is mainly related to a cross-shelf pressure gradient, which responds to an annual hydrological cycle. Current records from Shelikof Strait substantiate the presence of an annual signal and indicate that wind forcing has maximum effect from December through February, but it does not appear to augment flow at other times.

INTRODUCTION

Studies of circulation on the continental shelf have increased markedly during the last decade. Although shelf circulation patterns vary regionally, there are general classes of forcing mechanisms for shelf flow. As noted by *Smith* [1978], over an outer continental shelf region, dominant energy sources are atmospheric variability caused by passing storm systems and fluctuating ocean currents. On the shelf the main driving forces for 'first-order' flow are winds and tides [*Csanady,* 1976]. Seasonally varying wind stress and ensuing mass redistribution result in coastal flow off the Oregon coast [*Smith,* 1974; *Huyer et al.,* 1975; *Kundu and Allen,* 1976]. Over the southeast Bering Sea shelf, tides dominate horizontal kinetic energy [*Reed,* 1978; *Schumacher et al.,* 1979*a*]. Model results [*Semtner and Mintz,* 1977] and momentum considerations [*Csanady,* 1978; *Beardsley and Winant,* 1979] suggest that oceanic circulation results in the observed residual flow off the east coast of the United States, and runoff appears to be only locally important [*Beardsley and Hart,* 1978]. A model of winter circulation for the Adriatic Sea [*Hendershott and Rizzoli,* 1976], however, suggests that river runoff is a significant driving force, and a recent shelf circulation model [*Pietrafesa and Janowitz,* 1979] indicates that, in the absence of wind stress, a reasonable value of surface bouyancy flux results in alongshore velocities of about 20 cm/s.

Historically, little information is available on properties, flow, or dynamics of shelf waters in the northwest Gulf of Alaska. Most investigations were confined to studies of the offshore boundary current, the Alaskan Stream, and the subarctic gyre in general [*Dodimead et al.,* 1963; *Roden,* 1969; *Thomson,* 1972; *Favorite et al.,* 1976]. The Alaskan Stream is the northern boundary of the subarctic gyre and acts as a return flow for transport driven by wind-stress curl northward into the Gulf of Alaska [*Favorite et al.,* 1976]. An estimate of mean baroclinic transport for the stream off Kodiak Island is 12×10^6 m^3/s; however, variations do not reflect the large seasonal signal of wind-stress curl [*Reed et al.,* 1980].

Recent studies of coastal circulation in the northeast Gulf of Alaska [*Hayes and Schumacher,* 1976] suggest that, while oceanic forcing is dominant at the shelf edge, the inner shelf or coastal circulation differs from the shelf-break flow. *Royer* [1979] discussed the impact that extensive precipitation and runoff have on dynamic height and sea level along the Gulf of Alaska coast; at high latitudes, and consequent low temperature, salinity primarily controls density. Royer estimated that the maximum monthly rate of freshwater addition is about 20×10^3 m^3/s.

Preliminary results from part of the Outer Continental Shelf Environmental Assessment Program (OCSEAP) of the Bureau of Land Management and the National Oceanic and Atmospheric Administration indicate that in the northwest gulf, the Alaskan Stream is an offshore feature whose typically high velocities do not extend onto the shelf [*Schumacher et al.,* 1978, 1979*b*]. Their hydrographic data, however, indicate that there is a shoreward flux of heat and salt. In this paper we present further evidence that a well-defined coastal flow exists over the inner shelf. This flow, which we call the Kenai Current, has a transport up to 1×10^6 m^3/s toward the west and is primarily driven by the baroclinic component of a cross-shelf pressure gradient.

SETTING

This investigation is concerned with a region of the western Gulf of Alaska in the vicinity of the Kenai Peninsula, Kodiak Island, and Shelikof Strait (Figure 1). The area is characterized by extremely complex bottom topography shoreward of the 1830-m isobath, and the continental shelf is excised by several deep troughs which alternate with prominent banks. The northernmost of these features is Amatuli Trough, which is a broad, deep (>200 m) cleft in the shelf as indicated by the 183-m isobath. Water depths in excess of 150 m occur within 10 km of the Kenai Peninsula. Two passages, Kennedy and Stevenson entrances, enter Shelikof Strait between the Kenai Peninsula and Afognak Island. Depths in Kennedy Entrance are as great as 200 m, but east and west of this passage, sill depths are about 150 m. Stevenson Entrance has an average depth of about 120 m. The combined cross-sectional area for these passages is about 5×10^6 m^2. Shelikof Strait lies between the Alaska Peninsula to the northwest and Kodiak and Afognak islands to the southeast. To the north, the strait connects with lower Cook Inlet. Maximum depths in the upper strait are generally 175 m, and the cross-sectional area is about 6×10^6 m^2. To the south of Amatuli Trough lies Portlock

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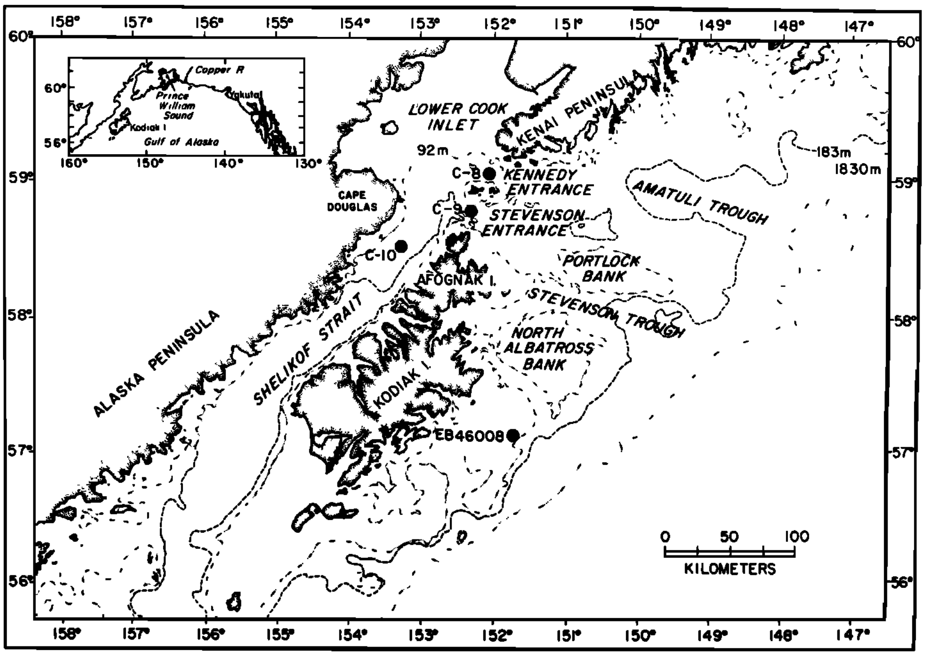

Fig. 1. Northwest Gulf of Alaska setting with prominent features and place names. The location of current meter stations and the NOAA data buoy are indicated.

Bank, which is an extensive region with depths between 45 and 60 m. The trough separating it from North Albatross Bank is indicated by a marked inshore trend of the 183-m isobath. The rugged undersea topography is similar to features above sea level; the coastline is ringed by coastal mountains with numerous peaks in excess of 3 km within 50 km of the shore. These features have major effects on winds and, as noted by *Royer* [1979], amplify precipitation effects on the Kenai Peninsula.

Data Acquisition and Processing

Conductivity and temperature versus depth (CTD) data were obtained during seven cruises (see Table 1) conducted by NOAA's Pacific Marine Environmental Laboratory from March 1977 through October 1978. CTD data were collected using Plessey model 9040 systems with model 8400 data loggers. These systems sampled five times per second for values of temperature, conductivity, and pressure. Data were recorded only during the down-cast using a lowering rate of 30 m/min. Nansen bottle samples were taken at most stations to provide temperature and salinity calibration. Data from monotonically increasing depth were 'despiked' to eliminate excessive values and were averaged over 1-m intervals to produce temperature and salinity values from which density and geopotential anomaly were computed.

Current meter station locations are shown in Figure 1. Aanderaa model RCM-4 current meters were used on taut wire moorings with an anchor and acoustic release at the bottom and a subsurface buoyant float about 2 m above the top current meter. The taut wire mooring tends to minimize surface wave-induced noise on the meter's rotor.

Two time series were produced from edited current observations using a Lanczos filter [cf. *Charnell and Krancus,* 1976]. The first series was filtered so that over 99% of the amplitude was passed at periods greater than 5 hours, 50% at 2.9 hours, and less than 0.5% at 2.0 hours. This time series was used to calculate sequences of 29-day tidal harmonic analysis of all the current records. The second series was filtered to remove most of the tidal energy; it passed 99% of the amplitude at periods over 55 hours, 50% at 35 hours, and less than 0.5% at periods less than 25 hours. This series was resampled at 6-hour intervals for use in examining subtidal circulation.

During summer 1978 a number of satellite-tracked drifting buoys were also deployed as part of the OCSEAP work. These drifters had 'windowshade' drogues centered at about 10 m, and they were interrogated several times a day by the Nimbus 6 satellite. The position data were processed by an objective computer routine to derive valid daily positions, and the data through September 1978 were reported by *Hansen* [1978].

Wind measurements from Middle Albatross Bank were provided by NOAA's data buoy EB 46008 (formerly EB 72) located at 57°06'N, 151°45'W (Figure 1), approximately 60 km offshore. This buoy was installed in August 1977 and provided data from November 1, 1977, until February 1979. Winds were sampled every 3 hours by satellite transmission link and were averaged for 8 min prior to transmission. Data gaps less than 6 hours were filled by linear interpolation, and results from a sea-level pressure analysis by Fleet Numerical Weather Central (U.S. Navy) were used for longer gaps.

Analysis of Hydrographic Data

Geopotential topography. Limited evidence was presented [*Schumacher et al.,* 1978] that net flow from current records

TABLE 1. Dates of Observations in the Vicinity of Kodiak Island and Shelikof Strait Used in This Study

Date	NOAA Ship
March 2–10, 1977	*Discoverer*
September 5–11, 1977	*Surveyor*
October 13–22, 1977	*Discoverer*
March 4–17, 1978	*Surveyor*
March 6–24, 1978	*Discoverer*
May 26 to June 7, 1978	*Discoverer*
October 3–22, 1978	*Discoverer*

agreed with that inferred from the 0/100-dbar geopotential topography. The paths of drogued buoys tracked by satellite also support the circulation patterns deduced from the geostrophic relation; *Royer et al.* [1979] found reasonable agreement between the two methods over the shelf east of this area, and summer 1978 drogued buoy data [*Hansen,* 1978] around Kodiak Island showed paths that support flow inferred from the 0/100-dbar geopotential topography. The geopotential patterns presented here are in good agreement with additional direct current measurements and with physical property distributions discussed later.

We summarize features of geopotential topography from *Schumacher et al.* [1978, 1979*b*] which are relevant to this study. During March, September, and October 1977, two consistent flow regimes existed: (1) the southwestward flowing Alaskan Stream over the slope accompanied by inshore counterflows east of Portlock Bank and (2) a well-defined coastal flow along the Kenai Peninsula. Weak, variable circulation occurred between the two more organized flows. During September and October a southward extension of the coastal flow appeared east of Afognak and Kodiak islands and may have been important in initiating a well-developed (relief of ~0.04 dyn m) gyrelike feature between Portlock and North Albatross banks. During October the coastal flow had much lower surface salinities than during other periods. Relief across the coastal flow was about 0.05 dyn m in March and September but increased threefold in October. In the vicinity of the entrances to Shelikof Strait, geostrophic patterns for April–May 1972 also suggested westward baroclinic flow [*Favorite and Ingraham,* 1977].

Two cruises (March 1978, NOAA ship *Discoverer*, and October 1978) had adequate coverage to examine concurrent conditions off the Kenai Peninsula and in Shelikof Strait. We present these data to depict coastal conditions. The 0/100-dbar geopotential topography in March 1978 is shown in Figure 2*a*. The coastal flow was present, and its relief was about 0.06 dyn m. It entered Shelikof Strait through both passages and turned south near Cape Douglas. Conditions in October 1978 (Figure 2*b*) showed a more clearly defined coastal flow like that in October 1977. The relief across the flow in October 1978 was about 0.20 dyn m, and a portion of the flow appeared to turn south near Afognak Island. The remainder of the coastal flow, however, entered Shelikof Strait through Kennedy Entrance and turned south as a well-defined flow near Cape Douglas. As suggested by *Muench et al.* [1978], this flow seems to be a permanent feature.

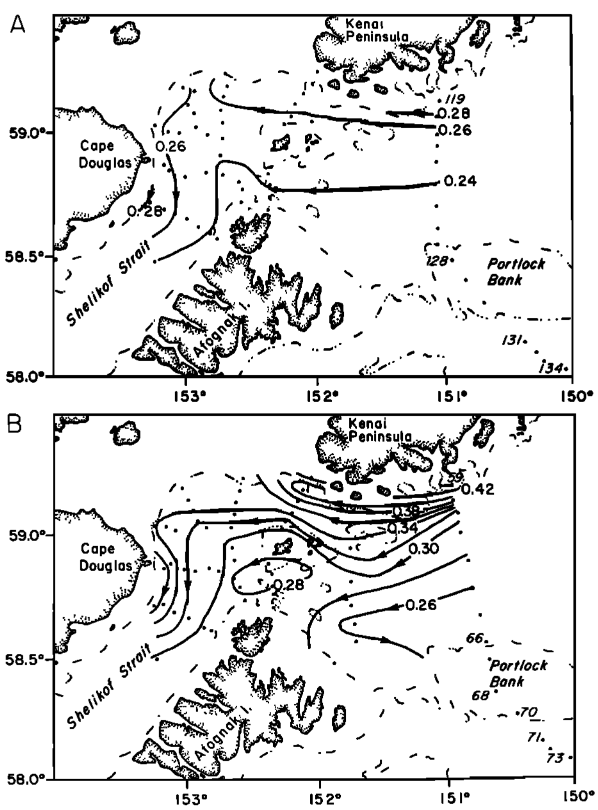

Fig. 2. Geopotential topography (0/100 dbar, in dyn m) observed during (*a*) March 13–21, 1978, and (*b*) October 9–22, 1978. Some CTD station locations are shown in italics.

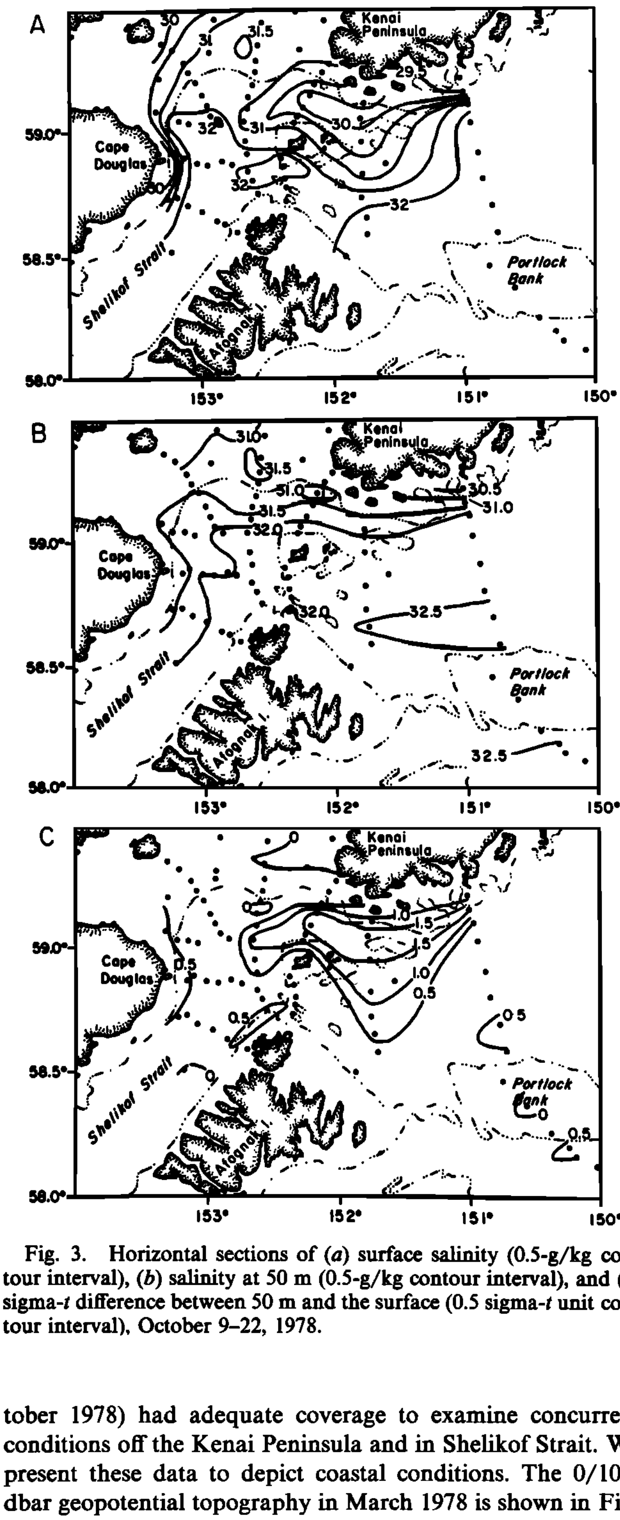

Fig. 3. Horizontal sections of (*a*) surface salinity (0.5-g/kg contour interval), (*b*) salinity at 50 m (0.5-g/kg contour interval), and (*c*) sigma-*t* difference between 50 m and the surface (0.5 sigma-*t* unit contour interval), October 9–22, 1978.

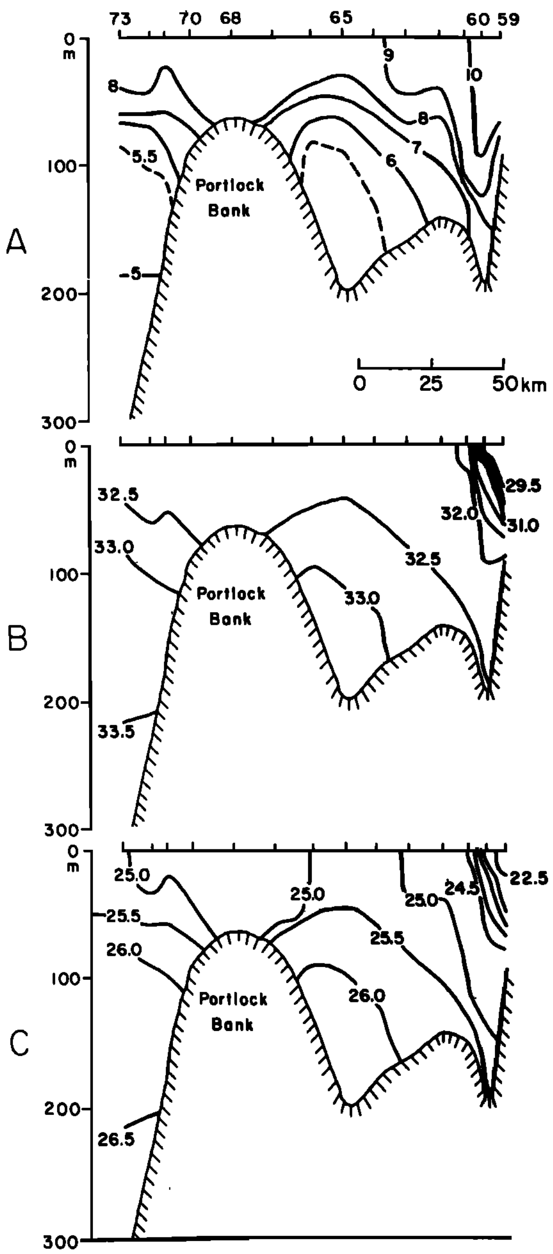

Fig. 4. Vertical sections of (*a*) temperature in degrees Celsius, (*b*) salinity in g/kg, and (*c*) sigma-*t*, October 10, 1978.

We emphasize that the coastal flow off the Kenai Peninsula is distinct from the Alaskan Stream. This is especially apparent from CTD data extending east of this area [*Royer et al.*, 1979], which we summarize as follows: the stream is present seaward of the shelf break, inshore there is very weak flow (often with reversals), but a continuous flow is present along the coast. The coastal flow had an alongshore extent of over 500 km, that is, east to the Copper River (near 145°W), and had high velocities and transport. We call it the Kenai Current. This term was first used by *Reed and Schumacher* [1979], who felt that the Kenai Peninsula was the land feature most clearly associated with the flow. Since this coastal flow does not appear to be well developed throughout the Gulf of Alaska, other names seem less appropriate. The Kenai Current is the dominant feature of shelf circulation in the northwest Gulf of Alaska with an alongshore extent of roughly 1000 km.

Physical property distributions. The distribution of surface salinity, salinity at 50 m, and the sigma-*t* difference between 50 m and the surface in October 1978 are shown in Figures 3*a*, 3*b*, and 3*c*, respectively. The surface salinity distribution is similar to the geopotential topography, but there are significant differences. Along the easternmost CTD section, low-salinity water (<32.0 g/kg) is confined quite close to shore and does not reflect the rather broad flow indicated in Figure 2*b*. The salinity pattern near Shelikof Strait is quite similar to the flow, even suggesting a counterflow on the north side of Kennedy Entrance as in Figure 2*b*. A new feature shown by surface salinity is an intrusion of low-salinity water from the north on the western side of Shelikof Strait; this could not be shown by the geopotential topography because most of the northwest side of Shelikof Strait is shoaler than the reference level (100 dbar). The salinity distribution at 50 m (Figure 3*b*) is also similar to the geostrophic flow pattern, and estuarine

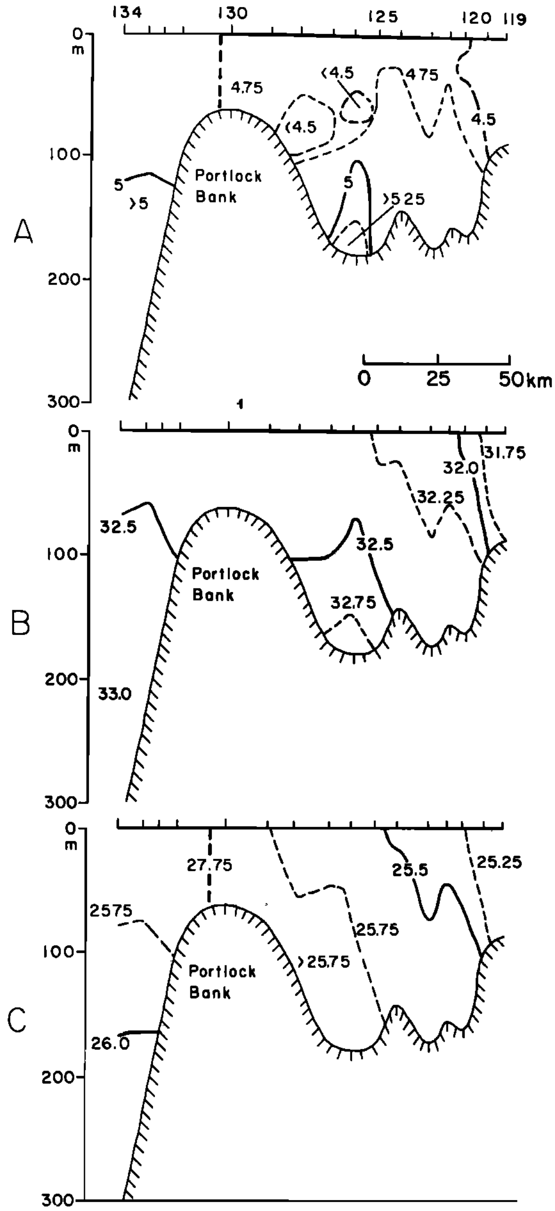

Fig. 5. Vertical sections of (*a*) temperature in degrees Celsius (*b*) salinity in g/kg, and (*c*) sigma-*t*, March 20, 1978. Additional contours have been included to emphasize features.

outflow (salinity < 31.0 g/kg) from Lower Cook Inlet is apparent as in the surface salinity distribution. The zone of high-salinity water outlined by the 32.5-g/kg isohaline south of the Kenai Peninsula coincides with the low in geopotential anomaly on the seaward side of the coastal flow.

The distribution of the sigma-*t* difference between 50 m and the surface (Figure 3*c*) indicates that greatest stratification is associated with the coastal flow. This is caused primarily by dilute (low-density) near-surface water, and values are reduced somewhat by enhanced tidal mixing through the entrances and near shore. Relatively large differences (>0.5) occur in waters off Cape Douglas, but the water column is isopycnal over Portlock Bank and over a shoal area stretching west of the Kenai Peninsula. This distribution supports geopotential features previously described and also depicts zones of vertical mixing which are strongly influenced by the effect of topography on tidal-power density per unit mass, V^3/H [*Fearnhead*, 1975], where V is mean tidal speed and H is water depth. Tidal mixing affects the lower part of the water column, and winds mix the upper layer; thus over banks, water may become isopycnal, especially when initial stratification and/or positive bouyancy flux are small.

The geopotential topography in March 1978 (Figure 2*a*) indicated an appreciably reduced flow into Shelikof Strait compared to October 1978. The horizontal patterns of topography during the two seasons were similar, however. The surface salinity distribution pattern in March 1978 (not shown) is similar to that in October 1978 (Figure 3*a*), but the lowest values in Shelikof Strait were about 1 g/kg greater in March than in October. Differences in the salinity at 50 m during the two seasons were less marked, however. The surface temperature distribution in March 1978 (not shown) shows the coldest water is associated with the southerly flow on the western side of Shelikof Strait, and waters in the Kenai Current are somewhat cooler than to the south. During October 1978, coastal waters at the surface and at 50 m were warmer than those in other regions. Coastal waters affected by land drainage tend to be relatively cooler in winter (but warmer in summer) than waters of oceanic origin because of the rapid heat transfer throughout the shallow water column.

Vertical sections of temperature, salinity, and sigma-*t* along the CTD sections off the Kenai Peninsula (stations 59–68 and 70–73 in October 1978, see Figure 2*b*; stations 119–134 in March 1978, see Figure 2*a*) are presented in Figures 4 and 5. The sections in October display a narrow band (~20 km) of warm, dilute, low-density water adjacent to the Kenai Peninsula; the strong slopes are in agreement with the westward coastal flow that enters Shelikof Strait. Indications of weak westward flow, however, extend well south of this band to the northern flank of Portlock Bank. The three stations (67, 68, and 70) over Portlock Bank exhibit weak stratification, and the water at station 68 (63-m depth) is isopycnal. South of Portlock Bank, the varying density slopes suggest that peak speeds in the Alaskan Stream were offshore of station 73. In March (Figure 5) the gradients across the coastal flow were much weaker than in fall, and there are no indications of significant baroclinicity elsewhere. Minimum temperatures were near the surface (rather than the bottom as in March), and everywhere there was less stratification than in fall. The isopycnal water atop Portlock Bank extended about 15 m deeper than in October, which suggests that mixing (presumably a combination of wind mixing near the surface and tidal mixing near the bottom if one ignores advective effects) was more efficient in winter than in summer through fall. Another feature shown by these data is that the intermediate and deeper parts of the water column were more saline in October than in March, while the coastal flow near the Kenai Peninsula was much more dilute during fall. This pattern of decreased surface salinity and increased intermediate and deep salinity during fall is consistent in our data set (Table 1); the results suggest that the saline water is 'drawn up' during periods of peak transport in the Kenai Current, perhaps as an interior upwelling in conjunction with buoyancy flux [*Pietrafesa and Janowitz*, 1979]. *Royer* [1975] proposed a similar process to explain the fall increase in deep salinity.

TABLE 2. Variations in Baroclinic Volume Transport and Maximum Surface Speed of the Westward Coastal Flow Off the Kenai Peninsula

Date	Volume Transport, 10^6 m^3/s	Maximum Speed, cm/s
March 4, 1977	0.4	15
September 10, 1977	0.4	30
October 19, 1977	1.0	89
March 6, 1978	0.3	14
May 29, 1978	0.1	13
October 10, 1978	1.2	133

Baroclinic transport and speed variations. The data presented thus far have indicated that coastal flow was more intense in October 1978 than during March of that year. The CTD section used to prepare Figures 4 and 5 was occupied on all of the Pacific Marine Environmental Laboratory (PMEL) cruises (Table 1). These data were used to compute baroclinic volume transport and maximum surface speed (both referred to the deepest common level) of the Kenai Current (Table 2). (A CTD section slightly west of this one was used for March 1977 because the former did not contain adequate data.) The data from four cruises in March, May, and September all indicate volume transport of 0.4×10^6 m^3/s or less and surface speeds not greater than 30 cm/s. In October of 1977 and 1978, however, transport was 1.0 and 1.2×10^6 m^3/s with maximum speeds of 89 and 133 cm/s. This marked increase in transport is in agreement with *Royer*'s [1979] conclusions that geopotential anomaly and alongshore flow respond to an annual hydrological cycle which has a fall maximum.

Precipitation often deviates considerably from the long-term mean, but these variations do not seem to have marked effects on the Kenai Current. Figure 6 shows the observed precipitation averaged for the National Weather Service stations at Cordova, Seward, and Kodiak during 1977 and 1978 and the long-term mean for thse stations [*National Oceanic and Atmospheric Administration*, 1977, 1978]. During 4 months prior to the March 1977 cruise, precipitation averaged over twice that of the long-term mean, but the March transport (0.4×10^6 m^3/s) and surface salinity do not seem to be unusual for the season. *Royer* [1979] concluded that accumulated land drainage over a large area was the major factor in producing variations in dynamic height. Hence monthly variations in precipitation, especially in winter when much of it is frozen, probably do not seriously alter the apparent seasonal pattern of large transport in fall and reduced values at other times. It seems likely, especially after considering *Royer*'s [1979] analysis of geopotential, sea level, and runoff, that the material presented in Figures 2–5 and Table 2 provides a reasonable estimate of mean conditions. Our use of data from only seven

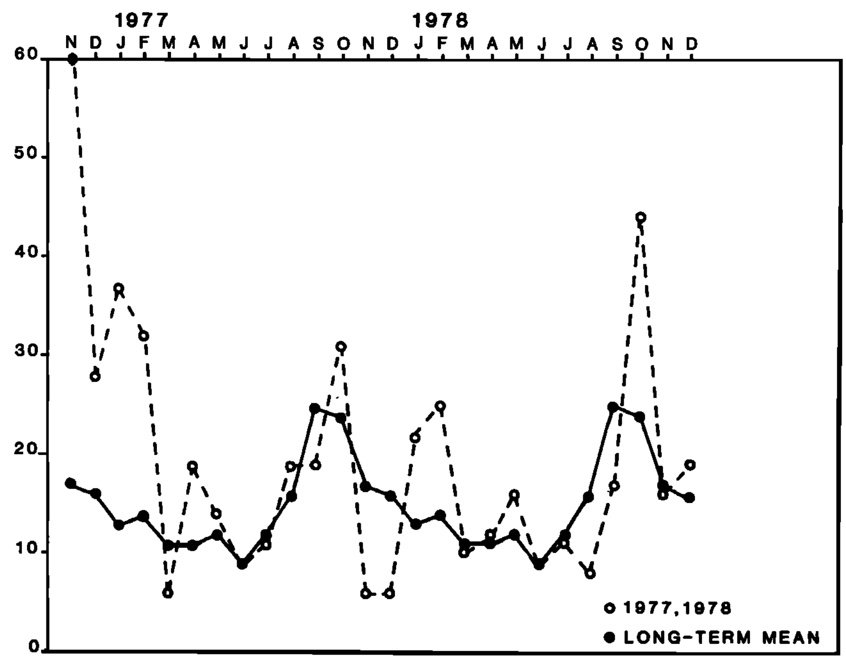

Fig. 6. Observed averaged precipitation (in centimeters) at Cordova, Seward, and Kodiak during 1977 and 1978 and the long-term mean.

cruises over a period of 19 months, of course, makes the interpretation of a seasonal signal somewhat tenuous; the very large and consistent variations in the data, however, and their coincidence in time with the mechanism discussed by Royer lends credence to our conclusion. Thus we expect the Kenai Current to normally have a transport of about 0.3×10^6 m^3/s and peak surface speeds of perhaps 20–30 cm/s; in early fall, however, transport rapidly increases to over 1×10^6 m^3/s, and peak surface speeds may exceed 100 cm/s. These extreme seasonal changes and very high velocities make the Kenai Current atypical of coastal flows [*Winant,* 1979] that are separate from major oceanic boundary currents.

ANALYSIS OF CURRENT OBSERVATIONS

Lagrangian measurements. Ten satellite-tracked drifting buoys were deployed between the Alaskan Stream and the Kenai Peninsula during May–July 1978 as part of OCSEAP studies [*Hansen,* 1978]. Two of these drifters were deployed within 40 km of shore and showed features of the Kenai Current; we summarize relevant trajectories from *Hansen* [1978] as follows. Drifter number 1473 moved westward in the southern part of the Kenai Current and turned south off Afognak Island as suggested by some of the property distributions and plots of geopotential topography (see, e.g., Figures 2*b* and 3*b*). Number 1775 moved west along the Kenai Peninsula and passed through Kennedy Entrance at net speeds of about 25 cm/s. It then moved eastward at very low speeds into Stevenson Entrance, where it eventually turned west and moved south through Shelikof Strait. In addition, a drifting buoy released as part of another project [*Reed,* 1979] entered the coastal flow at the head of the Gulf in December 1978, moved westward along the Kenai Peninsula into Shelikof Strait, and grounded on Cape Douglas in January 1979. Net speed for the 10-day portion of the track near the Kenai Peninsula was about 40 cm/s.

These data do provide direct evidence of an organized flow along the Kenai Peninsula with speeds comparable to those determined by the geostrophic relation. The westward movement into Shelikof Strait also agrees with the geopotential topography, but the apparent eastward movement out of Stevenson Entrance does not. We suspect that this may reflect a relatively brief flow 'event,' because similar features are suggested in some of the current meter records to be discussed.

Eulerian measurements. Current records were low-pass filtered and then averaged over 7 days on rotated coordinates which correspond approximately to the axes of the channels for Kennedy and Stevenson entrances and Shelikof Strait (see Figure 1). The coordinate systems are u positive 300°, v positive 030° for both entrances and u positive 225°, v positive 315° for Shelikof Strait. Wind stress was calculated following *Meyer et al.* [1979], where the drag coefficient is a function of wind speed for winds less than 15 m/s. Wind stress from buoy EB 46008 was resolved into alongshore (parallel to the Kenai Peninsula, u positive 240°) and cross-shelf (v positive 330°) components. The resulting net speeds through (toward the southwest) Shelikof Strait, net speeds into the entrances, and the alongshore and cross-shelf net wind stress are shown in Figure 7. Since the bulk of the Kenai Current flows through Shelikof Strait (Figure 2) and local contributions to flow are minimal [*Muench et al.,* 1978; *Schumacher et al.,* 1978], current records should be characteristic of flow along the coast.

During the winter observation period, flow in Shelikof Strait was always toward the southwest with typical speeds of 20–30 cm/s; during mid-October and early November, however, speeds were much higher, with a maximum of 70 cm/s. Similar characteristics were observed during the previous winter [*Schumacher et al.,* 1978]. During the peak flow period, vertical shear between the two meters at 25 and 72 m was about 5 times greater than the vertical shear over the remaining observation period. When baroclinic transport in the Kenai Current attains a maximum, baroclinic shear is not eliminated by mixing, although tidal speeds in Kennedy Entrance are strong (~60 cm/s). It is apparent that neither the alongshore nor cross-shelf wind stress was acting in a manner

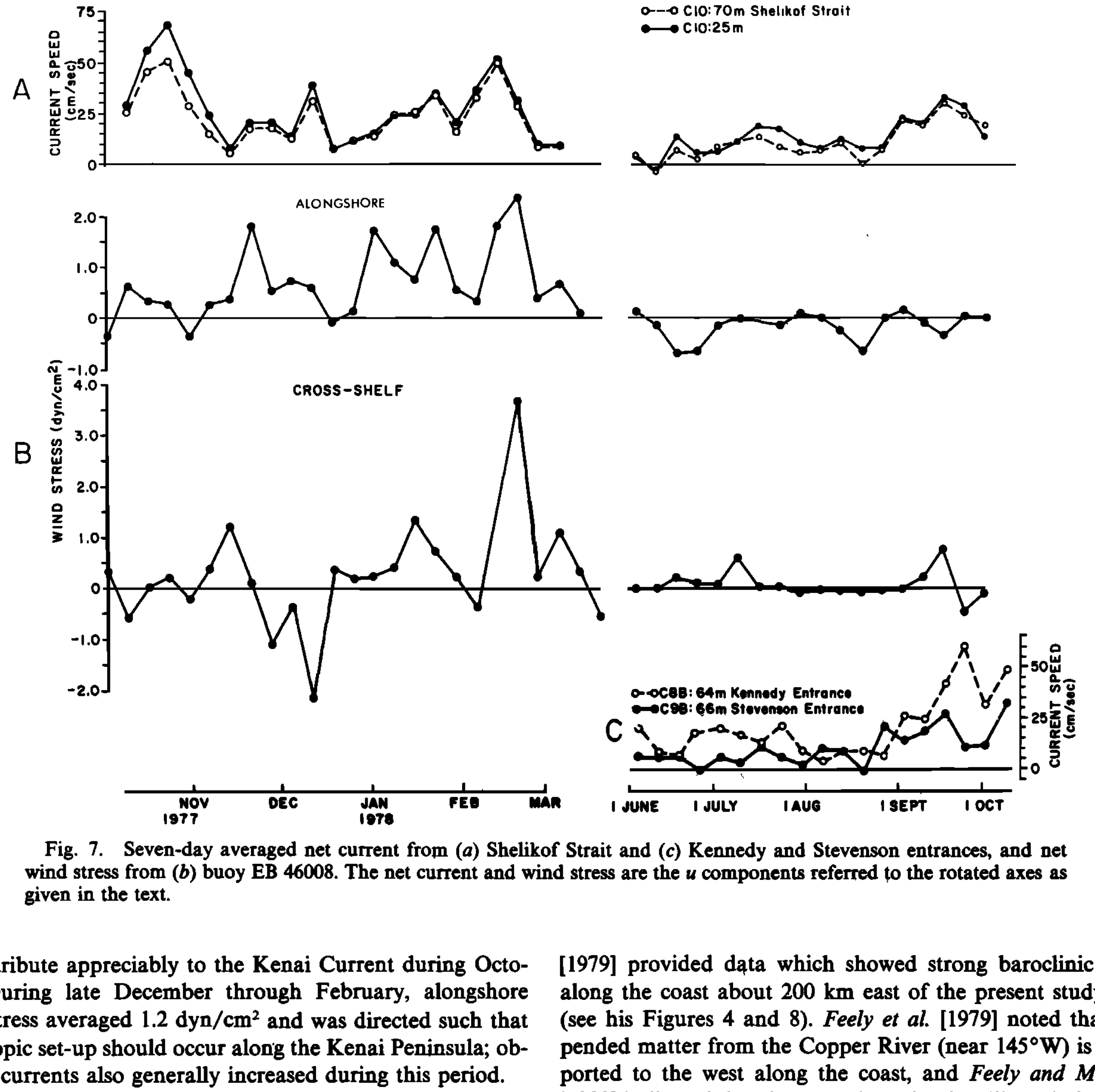

Fig. 7. Seven-day averaged net current from (*a*) Shelikof Strait and (*c*) Kennedy and Stevenson entrances, and net wind stress from (*b*) buoy EB 46008. The net current and wind stress are the *u* components referred to the rotated axes as given in the text.

to contribute appreciably to the Kenai Current during October. During late December through February, alongshore wind stress averaged 1.2 dyn/cm^2 and was directed such that barotropic set-up should occur along the Kenai Peninsula; observed currents also generally increased during this period.

Flow through Shelikof Strait over the summer observation period indicated a single flow reversal in early June. A few reversals were also noted in Stevenson Entrance but not in Kennedy Entrance. The records for Kennedy and Stevenson entrances suggest a pattern noted from the CTD data; flow is stronger and more persistent through Kennedy Entrance, and Stevenson Entrance, with its much weaker net flow, should be more responsive to forcing from wind events. Over the summer observation period though, neither alongshore nor cross-shelf wind stress is of proper direction or sufficient magnitude to significantly augment flow of the Kenai Current.

Discussion

Hydrographic and current observations presented above indicate that the Kenai Current is a continuous feature which extends for about 200 km along the Kenai Peninsula and in northern Shelikof Strait. Results from previous studies allow us to extend the horizontal scale of the Kenai Current. *Schumacher et al.* [1978] presented current and hydrographic data which indicated that the Kenai Current is evident throughout Shelikof Strait and probably follows depth contours across the shelf southwest of Kodiak Island. *Royer et al.* [1979] provided data which showed strong baroclinic relief along the coast about 200 km east of the present study area (see his Figures 4 and 8). *Feely et al.* [1979] noted that suspended matter from the Copper River (near 145°W) is transported to the west along the coast, and *Feely and Massoth* [1980] indicated that Copper River aluminosilicate is found in northern Shelikof Strait, which suggests that it flows there in the Kenai Current. Hence we conclude that the coastal flow (Kenai Current) exists from about 145°W along the Kenai Peninsula and through Shelikof Strait. This is a length of roughly 1000 km.

Although the impact of wind forcing on coastal flow cannot be quantified with the present data, wind observations support the concept of set-up due to alongshore wind stress during winter. Previous results [*Schumacher et al.*, 1979*b*] indicate that low-density (sigma-t < 24.0) surface waters off the Kenai Peninsula were compressed in horizontal extent from about 100 to 30 km in concert with increased alongshore wind. Thus winds can increase speeds in the Kenai Current by redistributing mass; however, the total alongshore baroclinic transport does not increase during this process. *Royer* [1979] showed strong correlation ($r = 0.93$) between dynamic height anomaly (0/200 dbar) and adjusted sea level at Seward and noted that the small difference between sea level and dynamic height cycles implied that annual changes in the barotropic current are small. The largest differences between sea level and dynamic height anomaly (see Royer's Figure 2) occurred in January through March, which is in agreement with our suggestion that barotropic set-up is important only from December through February.

We can estimate the magnitude of baroclinic and barotropic gradients using data from this and other shelf studies. The observed geopotential topography has a relief of 0.04–0.20 dyn m, which indicates an elevation of coastal sea level of 4–20 cm, over a cross-shelf length scale of about 25 km. The barotropic length scale, however, is likely to be larger than the baroclinic scale here, and the continental shelf width (~150 km) seems appropriate [*Hayes and Schumacher,* 1976]. Using March dynamic topography, as representing winter, the baroclinic gradient is about 10^{-6}. With a barotropic scale of 150 km and a 20-cm set-up (*Noble and Butman* [1979] give estimates) from the observed stress of 1.2 dyn/cm^2, the barotropic and baroclinic gradients of sea surface elevation might be approximately equal in winter. During summer and fall the baroclinic gradient increases while the barotropic gradient should be neglible because of the light winds (Figure 7).

Recent hypotheses for the residual flow off the east coast of the United States [*Csanady,* 1978; *Beardsley and Winant,* 1979] suggest that an alongshelf pressure gradient is generated by oceanic circulation, and a barotropic alongshelf flow results from this feature. The Alaskan Stream, however, is separated from the Kenai Current by a zone of very weak baroclinic flow or counterflow; hence it cannot be a major driving force for the Kenai Current. Water properties, of course, may be affected by transfer of stream water shoreward by eddies or an onshelf flux through the deep troughs. *Royer et al.* [1979] suggested that a combination of off-shelf Ekman flow in the upper layer and a cross-shelf pressure gradient (positive shoreward) resulting from freshwater input along the coast generate a shoreward subsurface entrainment flow. This may be valid during summer but cannot apply under typical winter wind stress conditions (set-up). Further experiments are required to determine the magnitude of the barotropic component of the Kenai Current and the mechanisms involved in generating the observed on-shelf flux of shelf-edge waters.

CONCLUSIONS

The organized coastal flow in the northwest Gulf of Alaska is a continuous, westward flowing current, the Kenai Current, which exists from the Copper River/Prince William Sound area through Shelikof Strait. This feature is clearly defined in geopotential topography as a narrow (15–30 km) band whose mean salinity is generally 0.5 g/kg less than that of adjacent shelf water. During October when the integrated effect of precipitation, river discharge, and melt water attains a maximum, the mean salinity of coastal flow is about 1.5 g/kg less than that of the seaward shelf waters. At such times, baroclinic transport is about 1.0×10^6 m^3/s. Baroclinic transport at other times is reduced, but still indicates a substantial (~0.3×10^6 m^3/s) westward flow. The magnitude of mean baroclinic transport (0.6×10^6 m^3/s) is about twice that observed over the Scotian shelf [*Drinkwater et al.,* 1979] or in the Adriatic Sea [*Hendershott and Rizzoli,* 1976], where cross-shelf density gradients resulting from freshwater addition also appear to be the main driving mechanism.

The Kenai Current is driven by a cross-shelf pressure gradient whose baroclinic component responds to an annual freshwater flux but does not 'spin-up' in winter during increased wind stress (Table 2). The barotropic pressure gradient component, however, seems to respond to seasonally varying wind stress; due to the characteristics of this forcing, the barotropic component augments flow only in winter. During winter, it may be of equal magnitude to the baroclinic gradient, but at other times it typically does not strengthen Kenai Current flow. Pressure gradients from oceanic currents do not seem to have a significant impact on the flow. This factor plays a major role in recent theories of coastal circulation, but our results suggest that 'coastal dynamics' cannot be unified so simply. The Kenai Current has an uncommon driving mechanism, and its speeds and transports are unusually large, particularly during the fall maximum.

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